Oxygen isotopes in fresh water biogenic opal - Northeastern US Alleröd-Younger Dryas temperature shift

Aldo Shemesh

Department of Environmental Sciences, Weizmann Institute of Science, Rehovot, Israel

Dorothy Peteet

NASA/Goddard Institute for Space Studies, New York, NY and Lamont Doherty Earth Observatory of Columbia University, Palisades, NY

Abstract. The first oxygen isotope analysis of biogenic opal from lake sediments, from the Alleröd/Younger Dryas transition in a core from Linsley Pond, Connecticut, gives an average estimate of a 6°C drop in temperature during the Younger Dryas. This shift represents temperatures during the bloom season, and may be less than the winter temperature drop. The sharp transition itself, with a duration of about 200 years, suggests that the temperature decrease may have been as large as 12°C. Previous estimates of the Alleröd/Younger Dryas temperature shifts are controversial, and range from 3-20°C, suggesting that further interdisciplinary research on the same samples is warranted.

One way that global climate change manifests itself is by redistributing energy throughout the globe. The Northern Hemisphere latitudinal temperature gradient during the late-glacial is at present a controversial topic. The magnitude of air temperature shifts during the Alleröd/Younger Dryas (YD) oscillation are estimated from mid-latitude pollen records surrounding the North Atlantic to be 3-5°C in Europe [Lowe et al., 1994] and 3-4°C in the eastern US [Peteet et al., 1993]. In contrast, lake temperatures estimates derived from aquatic midge larvae in the Canadian eastern maritimes and Maine range from 6-20°C, with larger shifts at more southern sites [Levesque et al., 1997]. The magnitude of YD cooling in Greenland ice cores ranges from at least 7°C from the Bölling warming [Dansgaard et al., 1989] to 15°C - a more recent estimate from borehole temperatures [Cuffey et al., 1995]. The ice core geochemical records reveal that massive frequent and short-term (decadal or less) changes in atmospheric composition occurred throughout this event, suggesting a very dynamic circulation [Mayewski et al., 1993].

To date, studies of oxygen isotopes in lake sediments have been restricted to the carbonate phase [McKenzie and Hollander, 1993], analyzing authigenic carbonates and calcareous bottom dwellers such as ostracods. Previous isotopic estimates of climate change from North America carbonates in lakes are few [Stuiver, 1970; Yu and Eicher, 1995] and although a late-glacial shift is apparent, it is undated. In Europe, the oxygen isotope estimates in the Swiss lakes range from 1-3 permil and were interpreted as a change in moisture source and air temperature [Eicher and Siegenthaler, 1976; Siegenthaler et al., 1984]. Because of the uncertainty in defining the moisture sources and transport pathways, the estimates of temperature change are also variable. Carbonate material suitable for isotopic analysis is not available in all lakes. However, many lakes contain diatomaceous opal in their sediment that can be used to retrieve climatic information.

We utilize, for the first time, the oxygen isotopic composition from biogenic silica, at Linsley Pond, Conn., eastern US, to investigate the temperature change associated with the Alleröd/YD oscillation. Linsley Pond (41°18'N, 72°45' W) is located in the town of North Branford about 17 km east of New Haven, Connecticut (Figure 1). To the west, it is bordered by a bedrock escarpment and to the south side by a swamp. Two small streams draining uplands enter the pond. Linsley Pond has a long history of limnological investigation, including its chemistry, plant and animal remains [Cowgill, 1970] and recent high-resolution investigations of major climatic [Peteet et al., 1993] and human [Brugam, 1978] disturbances. We selected Linsley Pond because previous pollen and macrofossil studies [Peteet et al., 1993] show a high sedimentation rate (1 meter per thousand years) for the late-glacial interval and a clear regional YD vegetational signal in the dramatic rise in alder, paper birch, spruce, and fir.

General characteristics of the climate of Connecticut include large temperature ranges, both daily and annual, equable distribution of precipitation among the four seasons, and considerable diversity of weather over short periods of time [NOAA, 1983]. Three types of air affect the state: cold, dry subarctic air; moist, warm air from the Gulf of Mexico and subtropical Atlantic; and cool, damp air from the North
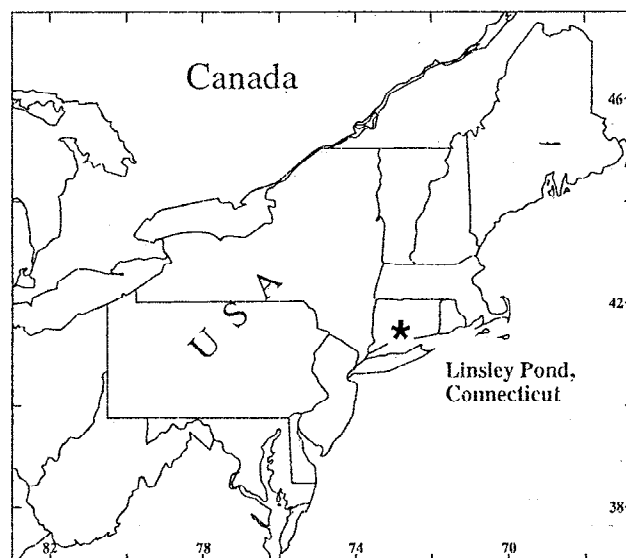
Figure 1. Location map.

Atlantic. Because the flow is usually from continental areas, Connecticut is more influenced by the first two types. A large number of storm centers pass near or over Connecticut throughout the year. Mean annual temperature along the Connecticut coast is 11°C, and average Jan. temp is about -6°C. Maximum temperatures in July are 29°C, and temperatures above 32°C occur only about 3 days per year. In coastal areas, about 190 days elapse between the last spring and first fall freeze. Annual precipitation averages 110-120 cm over most of the state. Measureable precipitation falls on an average of 1 day in 3, with a yearly total of 120 days. 9 to 22 centimeters fall in most months, while February and October are relatively dry with typically 7.5 centimeters. Annual average snowfall ranges from 155-220 cm, and most of the snow falls in January and February. During the colder months the prevailing wind is north to northwest over Connecticut, while from April to September southwest or south winds dominate. Mean hourly speed ranges from 11 km/hr in the summer and early fall to about 16 km/hour in the winter and spring. The sea breeze along the coast penetrates inland 8-16 km. "Northeasters" produce the greatest snowstorms in winter, with strong winds and heavy snow.

Over fifty years ago, a transect of sediment cores across Linsley Pond was used for diatom analysis and correlation [Patrick, 1943]. Lowermost meters of the lake sediment contained very few diatom species, consisting of *Cyclotella compta*, *Cyclotella ocellata*, *Fragillaria brevistriata*, *F. construens*, *Gomphonema acuminatum*, *Melosira ambigua*, *Rhopalodia gibba*, and *Synedra amphicephala*. As nutrients in the lake became more abundant, the flora became more diversified. (This sampling was not high-resolution enough to investigate oscillations such as the Alleröd/Younger Dryas). The early oligotrophic nature of the lake changed to eutrophic conditions around 10,000 yr BP (the C-1 palynological stage [Deevey, 1939]). Once eutrophic, the lake remained so until the present. No major changes in pH were seen throughout. Plankton counts in the lake (April 1937-June 1938) showed *Fragilaria crotonensis* (the most abundant diatom) reaching 4000 individuals per cubic centimeter during late July or August, while the next most abundant was *Synedra acus* with an average of 160 individuals per cubic centimeter, reaching maximum development from February to May.

For this new study, Linsley Pond samples were selected from the same 1987 12-m sediment core analyzed for pollen, macrofossils, and AMS stratigraphy [Peteet et al., 1993]. We analyzed 17 samples for opal at 5-cm intervals from 11.50 to 10.75 m depth. The age model (Table 1) utilizes the plant macrofossil AMS dates at 11.9-11.95 m (11,500 ± 300) and 10.8-10.85 m (10,440 ± 230) and assumes a linear sedimentation rate. The age resolution is approximately 50 radiocarbon years between samples.

Diatoms from Linsley Pond sediments were separated by differential settling and sieving combined with acid cleaning [Shemesh et al., 1992; Shemesh et al., 1995]. Diatoms measuring from 10 to 20 microns were cleaned and inspected microscopically to ensure the absence of contaminants. The isotopic exchange procedure for $\delta^{18}O_{si}$ analyses involved controlled isotopic exchange and fluorination [Juillet-Leclerc and Labeyrie, 1987]. Fluorination was performed at the Weizmann Institute, and the results are calibrated versus the NBS-28 standard with a long-term reproducibility of 0.14 per mil for diatom samples. The results are reported in the δ notation versus SMOW. Previous diatom analysis [Patrick, 1943] indicates that although diatom abundance varies throughout the core, many of the same species are present throughout the equivalent of the Alleröd/YD oscillation. The diatom assemblages include both planktonic and littoral species throughout. There is no indication of a major shift in species assemblages from the Alleröd zone to the YD zone.

Table 1. Age Model, Sample Depths, Diatom $\delta^{18}O_{si}$ (permil vs. SMOW) and Alnus (alder) Pollen Influx ($\times$ 10²/cm²/yr) from 1987 Linsley Pond core.

Depth (m)	Age (¹⁴C years BP)	Diatom $\delta^{18}O_{si}$	Alnus Influx
10.600	10,214		17
10.700	10,312		27
10.775	10,353	32.36	
10.800	10,410		19
10.825	10,440	30.73	
10.875	10,488	31.40	
10.900	10,508		52
10.925	10,537	32.77	
10.950	10,557		22
10.975	10,585	30.93	
11.000	10,605		25
11.025	10,634	32.55	
11.050	10,654		84
11.075	10,682	33.24	
11.100	10,703		47
11.125	10,730	32.56	
11.175	10,779	30.46	
11.200	10,801		4
11.225	10,827	30.71	
11.275	10,876	27.85	
11.300	10,899		6
11.325	10,924	26.29	
11.375	10,972	24.78	
11.400	10,997		10
11.425	11,021	28.27	
11.475	11,069	27.90	
11.500	11,094		5
11.525	11,118	26.77	
11.575	11,166	29.41	
11.600	11,192		0
11.700	11,290		1
11.800	11,388		1
11.900	11,485		1
12.000	11,583		1

Biogenic silica in lake sediments is composed primarily of diatoms (photosynthetic algae) that deposit internal opal ($SiO_2.nH_2O$) frustules. Silicification occurs only in the uppermost layer of the lake, because of light requirements for photosynthesis. Therefore, diatoms represent an ideal recorder of surface temperature and the isotopic composition of surface water, provided they deposit their silica with a known isotopic fractionation. Empirical calibrations [Shemesh et al., 1992; Gat, 1996] have shown that the oxygen isotopic composition of diatom opal ($\delta^{18}O_{si}$) depends on the temperature and the isotopic composition of the water ($\delta^{18}O_w$) from which it was deposited; the relation is given by:

$$t\ (^{\circ}C) = 11.03 - 2.03(\delta^{18}O_{si} - \delta^{18}O_w - 40) \qquad (1)$$

where t is the water temperature and $\delta^{18}O_{si}$ and $\delta^{18}O_w$ are the isotopic compositions of the biogenic opal and the lake water, respectively.

Resulting $\delta^{18}O_{si}$ measurements from core 1987 clearly show a sharp transition between Alleröd and the YD values (Figure 2, Table 1), followed by an abrupt rise in Alnus (Alder) pollen influx during the YD. Alleröd samples, older than 11,000 ¹⁴C, have an average $\delta^{18}O_{si}$ of 28.07 (±1.0) permil while YD samples, younger than 10,700 ¹⁴C, have an average $\delta^{18}O_{si}$ of 32.07 (±0.9). The transition occurs quickly, in about 200 years, and exhibits an isotopic shift of about 7 permil during this period. The average 4 permil isotopic enrichment of full YD conditions relative to Alleröd values is
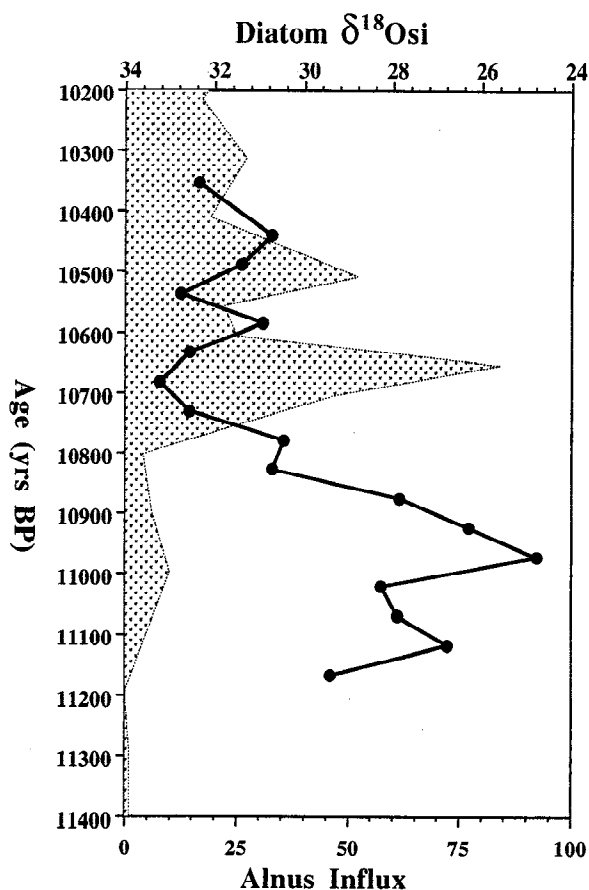
Figure 2. Diatom opal $\delta^{18}O_{si}$ (solid line, permil vs. SMOW) values *Alnus* (alder) pollen influx (shaded area) plotted against ^{14}C age from the 1987 Linsley Pond, Connecticut sediment core. $\delta^{18}O_{si}$ shift along with Alnus pollen influx increases indicate a temperature decline across the Alleröd/Younger Dryas boundary.

much larger than the variability within the YD and the Alleröd $\delta^{18}O_{si}$ The transition of $\delta^{18}O_{si}$ from Alleröd values to YD values can be explained by changes in surface water temperatures of the lake, changes in the isotopic composition of the lake water or any combination of both (eq. 1). An average 4 permil shift corresponds to a surface water cooling of 8°C during the YD, assuming that the lake water remained isotopically unchanged. As one can not rule out the possibility that lake water had changed its isotopic composition due to variation in the hydrological conditions, we conducted sensitivity tests in order to estimate the magnitude of the change. Under hydrologic steady state conditions the isotopic buildup in a lake is given by:

$$\delta^{18}O_w = \delta^{18}O_{in} + (\delta^{18}O_a - \delta^{18}O_{in} + \varepsilon/h)/(1 + F_{in}/E*(1 - h)/h)$$
(2)

where $\delta^{18}O_w$, $\delta^{18}O_{in}$ and $\delta^{18}O_a$ are the isotopic compositions of the lake water, input water and atmospheric moisture, respectively, h is the relative humidity (normalized to saturation at surface temperature), F_{in} and E are the water flux into the lake and the evaporative flux, respectively, $\varepsilon = (\alpha - 1)*1000 + \Delta\varepsilon$ where α is the temperature dependent fractionation factor of water in the liquid and the vapor phase, and $\Delta\varepsilon = C_k (1-h)$ where C_k is the kinetic constant which depends on the conditions at the air - water interface and

equals 15 for ^{18}O [Gat, 1996]. Thus, a change in one or more of the parameters in equation 2 can yield a change in lake water $\delta^{18}O_w$ and as a consequence a change in diatom $\delta^{18}O_{si}$. The most plausible factors that can change during a climatic shift are air temperature, h, F_{in}/E and $\delta^{18}O_{in}$. The term $(\delta^{18}O_a - \delta^{18}O_{in})$ is almost constant due to the mechanism of rain formation and does not change as long as precipitation is coupled to evaporation.

A 10°C warming in air temperature, from 10 to 20°C, results in 0.4 permil shift in $\delta^{18}O_w$ if the water content of the atmosphere (e_a) and the F_{in}/E are held constant. At an air temperature of 15°C and constant F_{in}/E, increasing the relative humidity from 0.4 to 0.8 yields an isotopic shift in $\delta^{18}O_w$ of 0.27 permil. Decreasing F_{in}/E from 8 to 5 yields a $\delta^{18}O_w$ change of 0.4 permil. Obviously, these parameters do not change independently in nature. Some of the effects tend to cancel out such as an increase in temperature that will cause an increase in evaporation resulting in a net isotopic effect on the lake water that is minimized. Thus, large changes in air temperature, humidity and inflow to evaporative fluxes do not significantly change the isotopic composition of the lake.

The sole parameter that can change $\delta^{18}O_w$ significantly is $\delta^{18}O_{in}$. There is no evidence that the local geology changed during the YD to produce a new upstream lake that would provide Linsley pond with evaporated and $\delta^{18}O$ enriched water. Therefore, a change in $\delta^{18}O_{in}$ can be obtained through a regional change in the hydrology and, more specifically, in the precipitation pattern of the area during the YD. At present, the precipitation in the area has two main sources of moisture. The first is derived from Atlantic air masses arriving from the Southeast and the second from continental air masses that arrive from the Northwest. Each source has a distinct isotopic signature. The Atlantic air masses carry precipitation with $\delta^{18}O$ of -3 to -5 permil while the continental system carries depleted values of -14 to -20 permil [Stuiver, 1970]. Thus, an increase in the relative contribution of Atlantic precipitation to the hydrological cycle during the YD will cause an isotopic enrichment of $\delta^{18}O_{in}$ and as a consequence, of $\delta^{18}O_w$. It is difficult to estimate the magnitude of this change as we do not have direct measurements on the relative strength of the Atlantic/Continental input to the rain. However, some constraints can be applied using the modern climate and hydrography of the lake. The $\delta^{18}O$ seasonal variability of surface and deep lake water are on the order of 2 permil, much smaller than the 10 permil observed in precipitation $\delta^{18}O$ at the same period [Stuiver, 1970]. It indicates that the largest changes in $\delta^{18}O$ precipitation are damped by the hydrological cycle. A first order isotopic mass balance calculation of the lake suggests that its modern value (-8 permil) is the result of mixing of 60 percent Atlantic rain and 40 percent continental rain. This calculation provides the upper limit for continental rain input as it ignores the effect of snow contribution to the hydrological cycle. If we further assume that snow is depleted relative to local rain and that snow provides about 1/10 of the annual precipitation, we calculate that the continental input to the precipitation balance is lower than 30 percent. Assuming that the continental input was reduced during the YD but did not completely cease [Rind et al.,1986; Mikolajewicz, 1996] we obtain a change in the hydrological input which on the order of 1 permil. Thus, we argue that only a small fraction, probably on the order of 1 permil, of the 4 permil shift between the Alleröd and the YD can be attributed to changes in local hydrology. We attribute the other 3 permil to a surface water cooling of about 6°C during the YD. We note that this

temperature change corresponds to the bloom season and that winter temperatures during the YD might have been colder than Alleröd winters by more than 6°C. We estimate that during the sharp transition period (approx. 10,950 - 10,780), with a duration of about 200 years, the surface water temperature drop might have been larger, about 12°C, corresponding to the 7 permil shift. These are also our best estimate for the change in air temperature as both lake temperature and local air temperature are strongly linked and approximate closely the large-scale air temperature (Muller-Navarra et al., 1997). The time for vegetation to reach equilibrium in the northeastern US is longer than this shift, and hence the brevity of the large change would have limited the magnitude of the vegetational signal. However, alder and birch are typical colonizing species after forest disturbance, and along with spruce and fir rise indicate a decline in temperature.

This 6°C isotopic estimate of a temperature shift during the Alleröd/Younger Dryas is at least 2°C larger than the pollen/macrofossil estimate which was calculated from altitudinal tree limits in the Adirondacks [Peteet et al., 1993]. However, it is significantly smaller than the estimate of 20°C in Maine derived from chironomid data [Levesque et al., 1997]. Differences in these results indicate that more research is needed to refine the temperature estimates, and that future investigations should include pollen/macrofossil, chironomid, diatoms, and opal isotope measurements on the same high-resolution sample series in order to critically assess leads/lags in vegetation response to climate change.

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A. Shemesh, Department of Environmental Sciences and Energy Research, Weizmann Institute of Science, Rehovot 76100, Israel. (e-mail: cishemes@wiccmail.weizmann.ac.il)

D. Peteet, NASA/Goddard Institute for Space Studies, 2880 Broadway, New York, NY 10025 and Lamont Doherty Earth Observatory of Columbia University, Palisades, NY 10964. (e-mail: petect@giza.giss.nasa.gov)